

fragments were too low to induce a ground-shock condition sufficient for cratering. However, we do suggest that this body developed bow-shock peak overpressures and shock-initiated winds in the atmosphere near the ground sufficient to cause incipient disturbances on the venusian surface, depending on rheologic states of the target materials.

The 10 km-diameter body traveled from 45 km altitude down to the surface in 2 s with less than 2% deceleration and would have caused major atmospheric, impact-cratering, and ejecta effects. We estimate that the final crater would have been 150 to 200 km across but only a few kilometers deep, which we infer would be due in part to the high temperature, low strength, rapid relaxation rate of the target rock. This large body generated an enormous bow-shock wave with pressures in excess of 600 kbar that reached the ground surface and swept outward at over 10 km/s. It also evacuated a region over 40 km in diameter along the trajectory of the body in the atmosphere. This hot (~20,000–200,000 K), low-density region continued to expand and remained open for over several minutes after impact with substantial amounts of high-angle ejecta traveling through it to high altitudes. The effects of the 10-km body in Venus' environment are still under analysis, but several of the major aspects appear similar to the asteroid-impact results calculated for the Earth. The energetics of such large events tend to initially overwhelm both atmospheres.

In summary, our numerical simulations indicate that for weak, cometlike objects impacting into Venus' atmosphere at 20 km/s, a 100-m-diameter body completely disintegrates at high altitude and produces no ground surface effects. A 1000-m-diameter body also disintegrates but produces a strong bow-shock wave immediately followed by a dispersed band of vapor, melt (?), and fragments that reach the ground to produce strong shock pressures and radial winds in the near-ground atmosphere. We suggest that these conditions could induce incipient surface disruption and produce subtle surface features, depending on the rheologic states of the target materials. We suspect that the dark and light splotches that are several tens of kilometers in diameter seen in the Magellan SAR images [3–8], and believed to be impact induced, could have been generated by cometlike bodies a few kilometers in diameter. The bodies could be somewhat smaller if the splotches were produced by stronger asteroidlike bodies that fragmented lower in the atmosphere; these cases remain to be explored.

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THE EFFECTS OF VENUS' THERMAL STRUCTURE ON BUOYANT MAGMA ASCENT. S. E. H. Sakimoto and M. T. Zuber, Department of Earth and Planetary Sciences, The Johns Hopkins University, Baltimore MD 21228, USA.

The recent Magellan images have revealed a broad spatial distribution of surface volcanism on Venus [1]. Previous work in modeling the ascent of magma on both Venus and Earth [2–5] has indicated that the planetary thermal structure significantly influences the magmatic cooling rates and thus the amount of magma that can be transported to the surface before solidification. In order to understand which aspects of the thermal structure have the greatest

influence on the cooling of ascending magma, we have constructed magma cooling curves for both plutonic and crack buoyant ascent mechanisms, and evaluated the curves for variations in the planetary mantle temperature, thermal gradient curvature with depth, surface temperature gradient, and surface temperature. The planetary thermal structure is modeled as

$$\frac{T}{T_0} = 1 - \tau \left(1 - \frac{Z}{Z_0}\right)^n \quad (1)$$

where T is the temperature, T_0 is the source depth temperature, $\tau = 1 - (T_s/T_0)$ where T_s is the planetary surface temperature, Z is the depth, Z_0 is the source depth, and n is a constant that controls thermal gradient curvature with depth. Equation (1) is used both for mathematical convenience and flexibility, as well as its fit to the thermal gradients predicted by the cooling half-space models [6]. We assume a constant velocity buoyant ascent, body-averaged magma temperatures and properties, an initially crystal-free magma, and the same liquidus and solidus for both Venus and Earth.

The cooling model for the plutonic ascent has been described in detail in earlier publications [2–5], and is a low Reynolds number, high Peclet number problem of heat transfer through a thin thermal boundary layer around a sphere. The resulting plutonic cooling curves, which are dominated by the convective cooling terms and strongly influenced by the planetary thermal structure, are then expressed mathematically by

$$\frac{T}{T_0} = \frac{J}{J + \gamma} \left(1 - e^{-(J + \gamma)t}\right) + e^{-(J + \gamma)t} + \left(\frac{J}{J + \gamma}\right) \left(\frac{\tau}{[(J + \gamma)t_0]^n}\right) \left[e^{-(J + \gamma)t} (-1)^n n! - \sum_{p=0}^n \frac{(-1)^p n! [(J + \gamma)t]^{n-p}}{(n-p)!} \right] \quad (2)$$

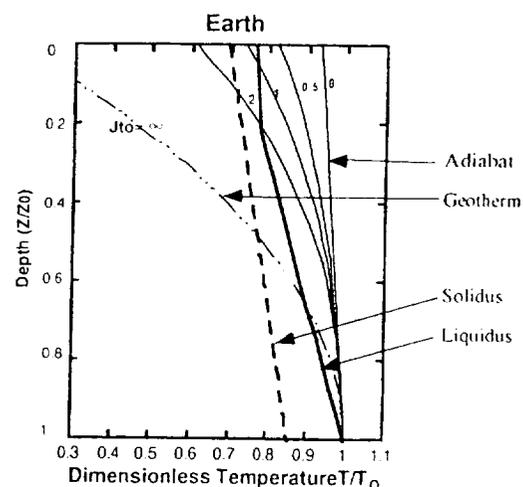


Fig. 1. Example of a typical cooling curve plot for pluton ascent. The curves are plotted for a given thermal structure and/or planet, and are contoured with their associated dimensionless ascent (Jt_0) values. In order for the magma to reach the surface unsolidified, the cooling curve must not cross the solidus. This plot is for a dry olivine tholeiite. The liquidus and solidus are obtained from [11]. Thermal structure parameters: $n = 2$; $dT/dZ = 1.0$ (dT/dZ of Earth); $Z_0 = 1.0$ (Z_0 of Earth).

where $J = 3NuK/a^2$, $\gamma = \alpha gV/C_p$, t is time, t_0 is total ascent time, $Nu = 0.8Pe^{1/2}$, $Pe = Va/K$, Pe is the Peclet number, V is the velocity of magmatic ascent, a is the body radius, $K (= 1 \times 10^{-6} \text{ m}^2\text{s}^{-1})$ is the thermal diffusivity, $\alpha (= 6 \times 10^{-5} \text{ deg}^{-1})$ is the coefficient of thermal expansion, g is the gravitational acceleration, and $C_p (= 1.25 \times 10^{-4} \text{ ergs Kg}^{-1}\text{K}^{-1})$ is specific heat capacity. T is the mean magma temperature, T_0 is the magma temperature in the source region, and n is a constant that defines the shape of the planetary thermal gradient (equation (1)). Equation (2) reduces to the expression for the thermal gradient (equation (1)) for an infinitely slow ascent (dimensionless ascent time $Jt_0 = \infty$), and to the adiabatic curve $T/T_0 = e^{-\gamma t}$ for an infinitely fast ascent ($Jt_0 = 0$). A typical plot of the resulting cooling curves for terrestrial conditions, contoured in Jt_0 values, is illustrated in Fig. 1. In order for the magma to reach the surface unsolidified, the cooling curve must not cross the solidus before it reaches the surface. The allowable Jt_0 values obtained from the cooling curve plots for Venus and Earth can be directly compared to obtain relative minimum magma ascent velocities, source depths, and body sizes. The results are shown in Fig. 2.

The cooling model for the buoyant crack ascent has previously been described briefly in [2]. It is the problem of a magma at an initial temperature T_0 placed in contact with the wall rock of temperature T_m . This problem was initially solved by [7], and their solution for the average temperature T as a function of time is

$$\frac{T - T_m}{T_0 - T_m} = \frac{8}{\pi^2} \sum_{m=1}^{\infty} \frac{\exp\{-(2m-1)^2 \pi^2 Kt/4a^2\}}{(2m-1)^2} \quad (3)$$

where the notation is the same as in equation (2), and the right side is constant for any single dimensionless ascent time (Kt/a^2). This result is for a constant wall rock temperature, but can be adapted to a variable wall rock (thermal gradient) temperature by approximating an incremental magma ascent in a simple numerical scheme where the initial magma temperature T_0 at any location m is, instead of the source depth temperature, the final magma temperature at the previous location $m-1$ [2]. If heat is conducted ahead of the magma body, the boundary temperature of the magma and wall rock will not be constant ($T_{\text{contact}} = T_m$) at any given location, but will be the average of the two initial temperatures of the magma and wall rock

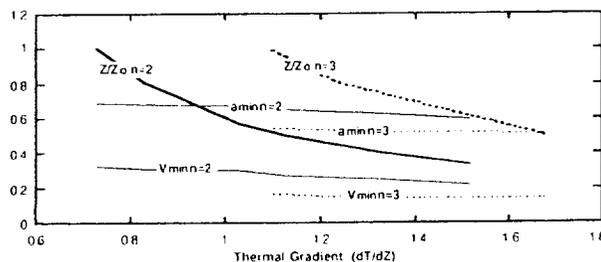


Fig. 2. Plot of the results for basaltic pluton ascent on Venus. The results are nondimensionalized by the corresponding terrestrial values, and the source depth values are the direct result of the assumed planetary thermal structure (equation (1)). For example, for a Venus thermal gradient the same as the terrestrial thermal gradient and $n = 2$, the corresponding minimum source depth is about 60% of the terrestrial value the minimum ascent velocity is about 30% of Earth's and the minimum body size is about 70% of Earth's. Thermal structure parameters: $n = 2$; $dT/dZ = 1.0$ (dT/dZ of Earth); $Z_0 = 1.0$ (Z_0 of Earth).

at any location for the majority of the cooling time [8]. For this case, the contact temperature at the m th position is the average of the final temperature of the magma at the $m-1$ position (T_{m-1}) and the initial wall rock temperature at the m th position (T_{m0}). The preliminary results from this model indicate that the effect of the planetary thermal structure is of the same order of the effect seen in the pluton model.

In general, for both ascent mechanism models presented here, the influence of the planetary thermal structure parameters for Venus in the probable order of decreasing importance is surface temperature, surface temperature gradient, thermal gradient curvature with depth, and planetary mantle temperature. The higher surface temperature of Venus, for otherwise similar planetary thermal structures, allows considerably smaller minimum possible crack sizes and/or magma body sizes, and slower ascent velocities than would be possible on Earth for a reasonable range of Venus source depths and surface thermal gradients. This surface temperature effect is greater for more primitive magma compositions, and may be greater for magmas of higher crystallinity. A higher venusian surface thermal gradient has the same effect of the higher surface temperature on magma transport, but to a much lesser degree. Similarly, for higher values of thermal gradient curvature with depth (higher n in equation (1)), the minimum possible ascent velocity and body/crack size also decreases slightly. If the mantle temperature for Venus is elevated by a hundred degrees or so over that of Earth [9], it should result in a modest increase of melt production and magma transport to the surface compared to Earth. The effect of the range of Venus surface temperatures with elevation ($390^\circ\text{--}470^\circ\text{C}$ or $660^\circ\text{--}740^\circ\text{K}$ [10]) is under investigation, and is also anticipated to have a significant effect on magma transport, possibly greater than that of the higher mantle temperature.

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EVIDENCE FOR RETROGRADE LITHOSPHERIC SUBDUCTION ON VENUS. David T. Sandwell¹ and Gerald Schubert², ¹Scripps Institution of Oceanography, La Jolla CA 92093, USA, ²Department of Earth and Space Sciences, University of California, Los Angeles CA 90024, USA.

Though there is no plate tectonics per se on Venus [1], recent Magellan radar images [2] and topographic profiles [3] of the planet suggest the occurrence of the plate tectonic processes of lithospheric subduction [4] and back-arc spreading [5]. The perimeters of several large coronae (e.g., Latona, Artemis, and Eithinoha) resemble Earth subduction zones in both their planform and topographic profile. McKenzie et al. [4] have compared the planform of arcuate structures in Eastern Aphrodite with subduction zones of the East Indies. The venusian structures have radii of curvature that are similar to those of terrestrial subduction zones. Moreover, the topography of the venusian ridge/trench structures is highly asymmetric with a ridge on the concave side and a trough on the convex side; Earth subduction zones generally display this same asymmetry.